THE UNIVERSITY OF CHICAGO
FOUNDED BY JOHN D. ROCKEFELLER

The Decennial Publications

A CONTRIBUTION TO THE THEORY OF GLACIAL MOTION

BY

T. C. CHAMBERLIN

The Decennial Publications

FIRST SERIES

Two volumes of Reports and eight volumes of Investigations, the latter consisting of a collection of articles representing the work of research of the several departments of the University. Size, quarto.

Volume I. The President's Report. A. Administration.

Volume II. The President's Report. B. Publications of the Members of the University.

Volume III. Part 1. — Systematic Theology, Church History, Practical Theology.
Part 2. — Philosophy, Education.

Volume IV. Political Economy, Political Science, History, Sociology.

Volume V. Semitic Languages and Literatures, Biblical and Patristic Greek, Comparative Religion.

Volume VI. Greek Language and Literature, Latin Language and Literature, Sanskrit and Indo-European Comparative Philology, Classical Archaeology.

Volume VII. Romance Languages and Literatures, Germanic Languages and Literatures, English, Literature in English.

Volume VIII. Astronomy and Astrophysics.

Volume IX. Mathematics, Physics, Chemistry, Geology.

Volume X. Zoölogy, Anatomy, Physiology, Neurology, Botany, Pathology, Bacteriology.

SECOND SERIES

A series of volumes embodying original research, consisting of systematic treatises, unpublished documents, and the like. Size, octavo.

THE UNIVERSITY OF CHICAGO
FOUNDED BY JOHN D. ROCKEFELLER

THE DECENNIAL PUBLICATIONS

A CONTRIBUTION TO THE THEORY OF GLACIAL MOTION

homas hrowder
BY
T. C. CHAMBERLIN
PROFESSOR AND HEAD OF THE DEPARTMENT OF GEOLOGY

PRINTED FROM VOLUME IX

CHICAGO
THE UNIVERSITY OF CHICAGO PRESS
1904

PRINTED MAY 1, 1904

A CONTRIBUTION TO THE THEORY OF GLACIAL MOTION

T. C. CHAMBERLIN

THE key to this study is the thesis that a glacier is a mass of crystalline rock—the purest and simplest type of crystalline rock known—since it is made up of a single mineral, of simple composition and rare purity, which never appears in a solid state except in the crystalline form.

THE GROWTH AND CONSTITUTION OF A GLACIER

The origin and history of a glacier is little more than the origin and aggregate history of the crystals that compose it. The fundamental conception of a glacier is therefore best obtained by tracing the growth of its constituent crystals. A basal fact ever to be kept in mind is that water in the solid form is always controlled by crystalline forces. When it solidifies from the vapor of the atmosphere, it takes the form of separate crystals (Plate I, Fig. 1). Perfect forms are developed only when the flakes fall quietly through a saturated atmosphere which allows them to grow as they descend. Under other conditions, the crystals are imperfect in growth and are mutilated by impact. But, however modified, they are always crystals. The molecules are arranged on the hexagonal plan, and the assemblage is controlled by a strong force, as the expansive power of freezing water shows. Once the definite crystalline arrangement is established, the molecules can be displaced only by the expenditure of a very notable amount of energy.

Snow crystals often continue to grow so long as they are in the atmosphere; but if they pass through an undersaturated stratum of air, or a stratum whose temperature is above 32° F., they suffer from evaporation or melting. When they reach the ground, the processes of growth and decadence continue, and the crystals grow or diminish according to circumstances.

A glacier is a colossal aggregation of crystals grown from snowflakes to granules of much greater sizes.

The microscopic study of new-fallen snow reveals the mode of change from flakes to granules. The slender points and angles of the former yield to melting and evaporation more than the more massive central portions, and this change probably illustrates a law of vital importance. It may often be seen that the water melted from the periphery of a flake gathers about its center, and if the temperature be right, it freezes there. This is a first step toward the pronounced granulation of snow, illustrations of which are familiar in snow that has lain long on the ground. If measured systematically from day to day, the larger granules taken from beneath the surface of coarse-grained snow are found to be growing. In a series of experiments[1] to determine the

[1] Carried out by Mr. C. E. Peet and Mr. E. C. Perisho, under the direction of the author.

law of growth, it was found that when the temperature of the atmosphere was above the melting-point, the growth was appreciably more rapid than when the air was colder, but that there was, on the average, *an increase under all conditions of temperature.* A portion of this average increase of the larger granules appears to come from the diminution and destruction of the smaller ones, for the total number of granules steadily diminishes. A portion of the growth doubtless comes from the moisture of the atmosphere which penetrates the snow, and another portion from the moisture derived from surface melting; but beneath the surface of a large body of snow the growth of the large granules probably takes place chiefly at the expense of the small ones. To follow the process, it should be noted that the free surface of every granule is constantly throwing off particles of vapor; that the rate at which the particles are thrown off is dependent, among other things, on the curvature of the surface being greater the sharper the curve; that the surfaces of the granules are at the same time liable to receive and retain molecules thrown from other granules; and that, other things being equal, the retention of particles also depends on the curvature of the surface, but in the reversed sense, the less curved surface retaining more than the sharply curved one. Under these laws, it is obvious that the larger granules of smaller surface curvature will lose less and gain more, on the average, than the smaller granules of greater curvature. It follows that the larger granules will grow at the expense of the smaller. It is also to be noted that, other things being equal, small granules melt more readily than large ones, and that where the temperature is nicely adjusted between melting and freezing the smaller may lose while the larger gain.

Another factor that enters into the process is that of pressure and tension. The granules are compressed at the points of contact, and put under tension at points not in contact; and the pressure and tension are, on the average, likely to be relatively greatest for the smallest granules. Tension increases the tendency to evaporation, and adds its effects to surface curvature. The capillary spaces adjoining the points of contact probably favor condensation. Ice expands in crystallizing, and pressure reduces the melting-point, while tension raises it. The effect of this is slight (0.0075° C. per atmosphere), and it probably plays little part in glacial action; but it is to be correlated with the much more important fact that *compression produces heat*, which may raise the temperature of the ice to the melting-point, while tension reduces the temperature to or below freezing. There is therefore a tendency for the ice to melt at the points of contact and compression, and for the water so produced to refreeze at adjacent points where the surface is under tension. This process becomes effective beneath a considerable body of snow, and here the granules gradually lose the spheroidal form assumed in the early stages of granulation, and become irregular polyhedrons interlocked into a more or less solid mass.

A third factor is also to be recognized, though its effectiveness is unknown. Under severe wind pressure, air penetrates porous bodies with appreciable facility. The "breathing" of soils and the curious phenomena of "blowing wells" and "blowing

caves" illustrate the effective penetration and extrusion of the air under variations of barometric pressure. In the snowfields, and in the more granular portions of glaciers near their heads, the porosity is doubtless sufficient to permit the appreciable penetration of the atmosphere. During a part of the time the probable effect is condensation of moisture from the air within the ice, and during another part, evaporation from the exposed surfaces of the granules within the ice. These alternating processes are attended by oscillations of temperature. While the balance between loss and gain of substance to the mass may be immaterial, the oscillating nature of the process and the fluctuations of temperature are probably favorable to the transference of molecules from point to point, and hence favorable to granular change.

Whether these processes furnish the total, or even the essential, explanation of the process or not, the observed fact is that there are all gradations from snowflakes and pellets into granular névé, and thence into glacier granules (*Gletscherkörner*), the last reaching the size of filberts and walnuts and beyond. The state of aggregation varies from the early, slightly coherent, granular stage, where the grains are small and spheroidal, and loosely attached, to the ice stage, where the cohesion has become strong through the interlocking growths of the large granules. Even when the mass has become seemingly solid ice, sufficient space is usually left between the granules to give a dispersive reflection to light which imparts to glacier ice its distinctive whitish color.

THE ARRANGEMENT OF THE CRYSTALLINE AXES

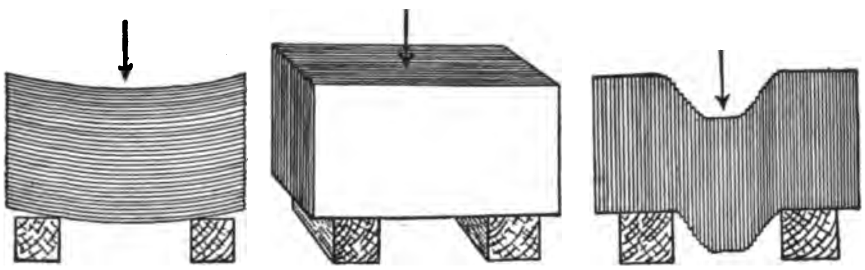
The most radical difference between glacier ice and ice formed directly from water is in the orientation of the crystals. In ice formed on undisturbed water the bases of the crystals are at the surface, and their principal axes are vertical, as shown by Mügge.[2] As they grow, the crystal prisms extend downward. This gives a columnar or prismatic structure to the ice, well shown when it is "honeycombed" by partial melting. In a glacier, on the other hand, the axes of the crystals, starting from snowflakes, lie originally in every direction, according to the accidents of their fall; and as the snow develops into ice, the principal axes of the crystals continue, in the main, to lie in every direction. Hence glacier ice, unlike pond ice, cannot usually be split along definite planes, except where cleavage planes are subsequently developed by extraneous agencies.

While the crystals of a glacier usually have their principal axes in various directions, there appears to be a tendency for them to approach parallelism in certain positions, especially in the lower part of a glacier near its terminus. Observations on this point are not so full and critical as could be desired, but it is probable that the parallel orientation is partly general and due to the vertical pressure of the ice, and partly special and local, and connected with the shearing planes and foliation (illustrated in Plate I, Figs. 2–4; Plate II, Figs. 5, 9–11; Plate III, Figs. 12–14).

The bearing of this partial parallelism of the crystals on shearing and foliation is

[2] MÜGGE, "Ueber die Plasticität der Eiskrystalle," *Neues Jahrbuch für Mineralogie*, etc., 1895, Bd. II, p. 211.

supposed to reside in the fact that a crystal of ice is made up of a series of plates arranged at right angles to the principal axis of the crystal. These plates may be likened to a pile of cards, the principal axis being represented by a line vertical to them. If a cube be cut from a large crystal of ice, it will behave much like a cube cut from the pile of cards. If the cube be so placed that its plates are horizontal, and if it be rested on supports at two edges and heavily weighted in the middle (Fig. 6), it will sag, the plates sliding slightly over one another so as to give oblique ends, but in this case the cube offers considerable resistance to deformation. If the cube be so placed that the plates stand on edge and stretch from support to support (Fig. 7), it will offer very great resistance to deformation; but if the plates be vertical and transverse to the line adjoining the supports (Fig. 8), the middle portion will sag under moderate weighting by the sliding of the plates on one another, and in a comparatively



6 7 8

FIGS. 6, 7, 8.—Diagrammatic figures to illustrate the method of deformation of crystals. (Adapted from Mügge.)

short time the middle portion may be pushed entirely out, dividing the cube. These properties have been demonstrated by McConnell[3] and Mügge,[4] and they appear to throw light on certain phases of the action of glaciers that are most pronounced in their basal parts and are best illustrated in arctic glaciers.

GLACIER TEMPERATURES

The temperature of glacier ice may range *downward* from the freezing-point of water, much as that of other solid portions of the earth's surface, but it has a fixed *upper* limit at 0° C., because all the heat it receives tending to raise its temperature above that point is converted into the latent form by the melting of the ice. The range of temperature is greatest at the surface, where it varies from 0° C. in the summer to the coldest temperature of the region where the ice occurs. Beneath the surface the range of temperature is more restricted, and increasingly so with increasing depth.

The variation of temperature at the surface is due primarily to the varying temperature of the air. During the cold season a wave of low temperature (the *winter* wave), starting at the surface, penetrates the ice, and during the warm season a wave of

[3] MCCONNELL, "On the Plasticity of Glaciers and Other Ice," *Proceedings of the Royal Society*, Vol. XLIV (1888), pp. 331-67 (with D. A. Kidd); Vol. XLVIII (1890), pp. 249-60; Vol. XLIX (1891), pp. 323-43.

[4] MÜGGE, *loc. cit.*

higher temperature (the *summer* wave) takes the same course. The day and night waves and other minor variables are, for present purposes, negligible.

The winter wave.—There are but few observations on the internal temperatures of glaciers during the winter season, but it seems certain that the winter wave diminishes rapidly downward and dies out below, much as does the winter wave which affects land surfaces not covered with ice. Conduction alone considered, the temperature of the ice where the winter wave dies out should correspond, approximately, to the mean annual temperature of the region, provided that temperature is below the melting-point of ice.

Assuming that in the high altitudes and high latitudes where glaciers abound the temperature of the surface may average about $-25°$ C. for the winter half of the year —which is about the case for north Greenland, Spitzbergen, and Franz Josef Land— and that the conductivity of the ice in the C. G. S. system is 0.005, the temperature would be lowered appreciably only about forty feet below the surface at the close of the period, conduction only being considered.[5] How far the internal temperature may be influenced by air forced through it by winds and by variations of the barometer is not known, and cannot well be estimated. The wave of low temperature descending from the surface in winter would probably become inappreciable before reaching a depth of sixty feet. At this depth the temperature should be about $-10°$ C., or near the mean annual temperature of the region.

The summer wave.—The summer wave in ice follows the analogy of the summer wave of ordinary earth much less closely, because of the melting of the ice. On this account, the summer wave is bifold. The one part travels downward by conduction; the other, by the descent of water. The one has to do primarily with the temperature before the melting-point of ice is reached; the other, with the temperature after that point is reached. The first conforms measurably to the warm wave affecting other solid earth-matter, while the second is governed by laws of its own. After the surface portion of the ice is brought to the melting temperature, the additional heat which it receives melts the ice and is transformed from *sensible* into *potential* heat. Ice charged with water is *potentially* but not *sensibly* warmer than ice which has just reached the melting temperature.

The warm wave of conduction dies out below, like the cold wave. The warm wave descending by the flow of water stops where the freezing temperature of water is reached. In regions where the average temperature is below freezing, the water wave does not descend so far as the wave of conduction, since the latter descends below the zone where the melting temperature is found.

The foregoing considerations warrant the generalization that glaciers normally consist of two zones: (1) an outer or upper zone of fluctuating temperature, and (2) an under zone of nearly constant temperature. The under zone does not exist where the thickness of the ice is less than the thickness of the zone of fluctuating temperature.

[5] Computation by E. D. K. Leffingwell.

This may be the case in very thin glaciers in very cold regions, and it may be true of the thin ends and edges of all glaciers.

The temperature of the bottom.—The internal heat of the earth is slowly conducted to the base of a glacier, where it melts the ice at the estimated average rate of about one-fourth of an inch per year. The temperature of melting is a little below 0° C., since prèssure lowers the melting-point at the rate of 0.0075° C. for one atmosphere of pressure. At the bottom of a mile of ice, therefore, the melting temperature is about −1° C. It is probable that in all thick glaciers the temperature of the bottom is constantly maintained at the melting-point. This may be indicated by the streams which issue from beneath glaciers during the winter, though the indication is hardly decisive, since the issuing waters may be derived partly or wholly from the rock beneath. In glaciers or parts of glaciers so thin as to lie wholly within the zone of fluctuating temperature, the temperature of the bottom is obviously not constant.

The internal temperature of the ice.—The range of temperature of the surface of a glacier has already been shown to lie between a maximum of 0° C. and the minimum temperature of the region where the glacier occurs. Lower, in the zone of fluctuating temperature, the range is less, and where the zone of fluctuating temperature passes into the zone of constant temperature, variation ceases. The thickness of the zone of fluctuating temperature varies with the temperature of the region where the glacier occurs, being greatest where the winters are coldest.

The range of temperature within the zone of constant temperature, in the case of all glaciers except thin ones in very cold regions, is from the mean annual temperature of the region at the top of the zone (provided this is not above the melting-point of ice) to the melting temperature of the ice at the bottom. Within these limits, the differences of temperature may be great or slight.

If we consider only the effects of the external seasonal temperatures and the internal heat of the earth, it appears that all the ice in the zone of constant temperature in the lower end of a typical alpine glacier should have a constant melting temperature; for the average temperature of regions where the ends of such glaciers occur is usually above 0° C., and this determines a temperature of 0° C. (or a little less) at the top of the zone, while a melting temperature is maintained at the bottom by the earth's interior heat. In thin glaciers of very cold regions, where the zone of constant temperature has relatively slight thickness, or lies wholly below the ice, the low temperature descending from the surface may so far overcome the effect of internal heat as to keep the bottom of the ice at a freezing temperature. In all other cases the ice at the bottom of the under zone has a melting temperature, while that above is probably colder.

In the higher altitudes and in the polar latitudes, where glaciers are chiefly generated, the mean annual temperature of the surface *is usually below the melting-point of ice.* Here the temperature of the ice between the top and the bottom of the zone of constant temperature must, on the average, be below the melting-point, unless

enough heat is generated in the interior of the ice to offset the effect of the temperature above. For example, where the mean annual temperature is $-6°$ C., as in certain high altitudes and high latitudes, the mean temperature in the zone of constant temperature should range from $-6°$ C. above to $0°$ C. (or a little less, because of pressure) below; *i. e.*, it should average about $-3°$ C. Under these conditions, all the ice in the zone of constant temperature, except that at its bottom, is permanently below the melting-point; but it is perhaps worthy of special note that much of it is but little below. In alpine glaciers the part of the ice affected by this constant freezing temperature is presumed to be chiefly that which lies beneath the snowfields. In polar glaciers the low temperature probably prevails beneath the surface not only throughout the great ice-caps, but also in the marginal glaciers which descend from them.

From these theoretical considerations we may deduce the generalization that in the zone of constant temperature *within the area of glacial growth* the temperature of the ice is generally *below* the melting-point, while *within the area of wastage* the temperature of the corresponding zone is generally *at* the melting-point.

Compression and friction as causes of heat.— The foregoing conclusions are somewhat modified by certain other sources of heat. The compression arising from gravity, and the friction developed where there is motion, are causes of heat. Since friction occurs only when motion takes place, the heat which it generates is secondary and may, for present purposes, be neglected. Compression not only lowers the melting-point slightly, but it *produces heat at the point of compression.* Where the ice is granular, the compression, due to weight, takes place at the contacts of the grains. At intermediate points the pressure tends to cause them to bulge, and this has the effect of lowering the temperature of the bulging areas. If therefore the compression be considerable, the granules may be warmed to the melting-point where they press each other, while at other points their temperature may be lower. In this case melting will take place at the points of compression, and the moisture so produced will be transferred to the adjacent parts of the granule and immediately refrozen. Melting at the points of compression would result in some yielding of the mass, and in some shifting of the pressure to new points where compression and melting would again take place. Thus the melting, the refreezing, and the attendant movement might go on until the limits of the power of gravity in this direction were reached. From considerations already adduced it would appear that the temperature in some parts of every considerable body of ice must be such as to permit these changes. This dynamic source of heat may modify the theoretical deductions drawn above from atmospheric and internal influences.

Summary.—If the foregoing generalizations be correct, (1) the surface of a glacier is likely to be melted during the summer; (2) its immediate bottom is slowly melting all the time, unless the thickness of the ice be less than the thickness of the zone of annual variation or of permanent freezing temperature; (3) its subsurface portion

in the zone of waste is generally melting, owing to descending water, compression, and friction; while (4) its subsurface portion *in the zone of growth* is probably below the melting-point, except as locally brought to that temperature by compression and friction, and at the bottom by conduction from the rock beneath.

FUNDAMENTAL AS DISTINGUISHED FROM AUXILIARY CAUSES OF MOTION

Since there must be motion in the area of growth to supply the loss in the area of waste, the *fundamental* cause of motion must probably be operative in bodies of ice the mean temperature of which is below the melting-point, *unless the internal sources of heat are considerable.* This fundamental cause does not exclude the *co-operation* of causes that work only (1) at the melting temperature, (2) where the ice is bathed with water, or (3) in the plane of contact between wet ice above and dry ice below. These may be *auxiliary* causes which abet the fundamental one in producing the more rapid movement of the warm season, or in bringing about especially rapid motion in situations where there is abundant water, or in inducing the shearing which is such a remarkable feature of arctic glaciers.

THE PROBABLE FUNDAMENTAL ELEMENT IN GLACIAL MOTION

Melting and refreezing.—It has just been pointed out that the *initial* or *fundamental* cause of glacial motion must be operative at the heads of glaciers where the temperature is lowest and the material most loosely granular. In this condition, there is reason to believe that motion takes place between the grains, rather than by their distortion through the displacement of their laminæ. The fact that the granular structure is not destroyed, as it would be by the indefinite sliding of the crystal plates over one another, sustains this view. The inference is that the gliding planes play a notable rôle in glacial movement only in the basal parts of the lower ends of glaciers, where the greatest thrusts are developed, and where the granules have become largest and most completely interlocked. At the heads of glaciers, where motion is initiated, there may be great downward pressure, but not vigorous thrusts from behind, and probably only moderate thrusts developed within the body itself. *There seems therefore no escape from the conclusion that the primal cause of glacial motion is one which may operate even under the relatively low temperature, the relatively dry condition, and the relatively granular texture which affect the heads of glaciers.*

These considerations lead to the view that movement takes place by the minute individual movements of the grains upon one another. While they are in the spheroidal form, as in the névé, this would not seem to be at all difficult. They may rotate and slide over one another as the weight of the snow increases; but as they become interlocked by growth, both rotation and sliding must apparently encounter more resistance. The amount of rotary motion required of an individual granule is, however, surprisingly small, and the meltings and refreezings incident to shifting pressures and tensions, and to the growth of the granules, seem adequate to meet the require-

ments. To account for a movement of three feet per day near the end of a glacier six miles long, the mean motion of the average granule relative to its neighbor is, roundly, one ten-thousandth of its own diameter per day, or one diameter in ten thousand days; in other words, it *changes its relations to its neighbors* to the extent of its diameter in about thirty years. A change of so great slowness under the conditions of granular alteration can scarcely be thought incredible, or even improbable, in spite of the interlocking which the granules may develop. The movement is supposed to be permitted chiefly by the temporary passage of minute portions of the granules into the fluid form at the points of greatest compression, the transfer of the moisture to adjoining points, and its resolidification. The points of greatest compression are obviously just hose whose yielding most promotes motion, and a successive yielding of the points that come in succession to oppose motion most (and thus to receive the greatest stresses) permits continuous motion. It is merely necessary to assume that the gravity of the accumulated mass is sufficient to produce the minute temporary liquefaction at the points of greatest stress, the result being accomplished, not so much by the lowering of the melting-point, as by the development of heat by pressure.

This conception of glacial "flowage" involves only the *momentary liquefaction of minute portions of the mass*, while the ice, as a whole, remains rigid, as its crystalline nature requires. Instead of assigning a slow viscous fluidity, like that of asphalt, to the *whole* mass, which seems inconsistent with its crystalline character, it assigns a free fluidity to a succession of particles that form only a minute fraction of the whole at any instant.

This conception is consistent with the retention of the granular condition of the ice, with the heterogeneous orientation of the crystals (in the main), with the rigidity and brittleness of the ice, and with its strictly crystalline character—a character which a viscous liquid does not possess, however much its high viscosity may make it resemble a rigid body.

Accumulated motion near the end of a glacier.—However slight the relative motion of one granule on its neighbor, the granules in any part of a glacier partake in the accumulated motion of the parts nearer the source, and hence all are thrust forward. Herein appears to lie the distinctive nature of glacial movement. Each part of a stream of water feels the hydrostatic pressure of neighboring parts (theoretically equal in *all* directions) and the momentum of motion, but not the rigid thrust of the mass behind. Lava streams are good types of viscous fluids flowing in masses comparable to those of glaciers, on similar slopes, and, in their last stages, at similar rates; but their special modes of flow and their effects on the sides and bottoms of their paths are radically different from those of glaciers. Forceful abrasion, and particularly the rigid holding of imbedded stones while they score and groove the rock beneath, are unknown in lava streams, and are scarcely conceivable. There is, so far as I know, neither experimental nor natural evidence that any typical viscous body, in flowing over a rugose bottom, picks up fragments and holds them as graving tools in its base

so fixedly as to cut deep long straight grooves in the hard bottom over which it flows. It would seem that competency to do this peculiar class of work, which is distinctive of glaciers, should be demonstrated before the viscous theory of glacial movement is accepted as even a provisional working hypothesis. Quite in contrast with viscous movement, it is conceived that a glacier is thrust forward rigidly by internal elongation, shears forcibly over its sides and bottom, and leaves its distinctive marks upon them.

AUXILIARY ELEMENTS IN GLACIER MOTION

Shearing.—In the lower portion of a glacier, where normally the thrusts are greatest, the granules fewest, and their interlocking most intimate, shearing appears to take place within the ice itself. This is illustrated by Figs. 2–5, and 9–14 (Plates I, II, and III). The shearing results in the foliation of the ice and in the forcing of débris between the sheared layers. Thus the ice becomes loaded in a special baso-englacial fashion (see particularly Fig. 14).

Within the zone of shearing, it is probable that the gliding planes of the crystals come into effective function. It is thought that the combined effect of the vertical pressure, the forward thrust, and the basal drag of the ice may be to increase the number of granules whose gliding planes are parallel to the glacier's bottom. At any rate, Drygalski reports[6] that there is a tendency to such an arrangement in the basal portion of the Greenland glaciers at their borders. It is conceived that where strong thrusts are brought to bear upon such a mass of granules, those whose gliding planes are parallel to the direction of thrust are strained with sufficient intensity to cause the plates to slide over one another, while those which are not parallel to the direction of thrust are either rotated into parallelism—when they also yield—or are pressed aside out of the plane of shear.[7] Shearing is observed to occur chiefly where the ice below the plane of shearing is protected more or less from the force of the thrust. It perhaps also occurs where the basal ice becomes so overloaded with débris that it is incapable of ready movement.

It is also probable that sharp differential strain and shearing are developed at the level where the surface water of the warm season descending into the ice reaches the zone of freezing. The expansion of the freezing water at the upper limit of the frozen zone may cause the layer affected by it to shear over that below. As the level of freezing is lowered with the advance of the warm season, the zone of shearing also sinks. This may be regarded as an auxiliary agency of shearing of application to the upper horizons.

[6] *Grönland-Expedition der Gesellschaft für Erdkunde zu Berlin*, 1891–93, Bd. I, pp. 491 ff.

[7] A series of experiments on compacted snow was conducted during a winter season to test this. About three-fourths of the granules in the tract of greatest strain took on a common orientation. The rigid resistance of the snow-ice proved very great, and differential pressures, increased to the breaking strength of the apparatus, nearly one thousand pounds per square inch, were insufficient to produce, in the time given, a definite shear-plane of the Greenland type, but the approximation attained lends strength to the belief that a new series of experiments, contemplated but as yet unexecuted, in which the rigidity of the snow-ice shall be fully anticipated and the time requisite for molecular change provided for, will fully sustain the view here entertained.

High temperature and water.—In the zone of waste, a higher temperature and more water lend their aid to the fundamental agencies of movement; and there is need for these aids to promote proportionate movement, for here the granules are more intimately interlocked, and the ice more compact, and inherently more solid and rigid. The average temperature here is, however, near the melting-point, and during the warm season the ice is bathed in water, so that the necessary changes in the crystals are facilitated, and movement apparently takes place even more readily than in the less compact granular portion of lower temperature and drier state. The extraordinary rate of movement of certain tongues of ice in some of the great fiords of Greenland is probably due to the convergence of very thick, slow-moving ice from the interior, into basins leading down to the fiords. Into the same basins a large amount of surface water is concentrated at the same time, with the result that the thick ice, bathed with water, and having a high gradient, develops unusual velocity during the warm season.

Application.—By a studious consideration of the co-operation of the auxiliary agencies with the fundamental ones, the peculiarities of glacial movement may apparently be explained. In regions of intense cold where a dry state and low temperature prevail, as in the heart of Greenland, the snow-ice mass may accumulate to extraordinary thickness, for the work of movement seems to be thrown almost wholly upon compression, with the slight aid of molecular changes due to internal evaporation and allied inefficient processes. Since the temperature in the upper part of the ice is very adverse, the compression must be great before it becomes effective in melting the contacts of the granules of ice, and hence the great thickness of the mass antecedent to much motion. Similar conditions more or less affect the heads of alpine glaciers, though here the high gradients favor motion with less thickness of ice; but in the lower reaches of alpine glaciers, where the temperatures are at or near the melting-point, and the ice is bathed in water, movement may take place in ice which is thin and relatively compact.

If the views here presented are correct, there is also near the end or edge of a glacier the co-operation of rigid thrust from behind with the tendency of the mass to move on its own account. The latter is controlled by gravity, and conforms in its results to the laws of liquid flow. The former is a derived factor, and is a mechanical thrust. This thrust is different from the pressure of the upper part of a liquid stream on the lower part, because it is transmitted through a body whose rigidity is effective, while the latter is transmitted on the hydrostatic principle of equal pressure in all directions.

CORROBORATIVE PHENOMENA

The conception of the glacier and its movement here presented explains some of the anomalies that otherwise seem paradoxical. While a glacier, in a sense, "flows" over a surface, it often cuts long deep furrows in firm rock (Plate III, Fig. 15). It is difficult to explain this if the ice be so yielding as to flow under its own weight on a surface which is almost flat. If the mass is really viscous, its hold on its imbedded

débris should also be viscous, and a bowlder in the bottom should be rotated in the yielding mass when its lower point catches on the rock beneath, instead of being firmly held while a deep groove is cut. This is the more to the point since viscous fluids flow by a partially rotatory movement. If, on the other hand, the ice is always a rigid body, which yields only as its interlocking granules change their form by loss and gain, a rigid hold on the imbedded rock at some times, and a yielding hold at others, is intelligible, for, on this view, the nature of its hold is dependent on the temperature and dryness of the ice. Stones in the base of a glacier may be held with very great rigidity when the ice is dry and cold, scoring the bottom with much force; while they may be rotated with relative ease when the ice is wet and warm. In short, the relation of the ice to the bowlders in its bottom varies radically according to its dryness and temperature. *A dry glacier is a rigid glacier. A dry glacier is necessarily cold, and a cold glacier is necessarily dry.*

On the view here presented, a glacier should be more rigid in winter than in summer, and the whole thickness of the glacier should experience this rigidity chiefly at the ends and edges, where the relative thinness of the ice permits the low temperature to reach its bottom. The motion in these parts during the winter is therefore very small.

In this view may also be found an explanation of the movement of glaciers for considerable distances on upward slopes, even when the *surface*, as well as the base, is inclined backward. So far does this go that superglacial streams sometimes run for some distance *backward—i. e.*, toward the heads of the glaciers—while in other places surface waters are collected into ponds and lakelets. Such a forward ascent of the surface of ice is not difficult to understand, if the movement be due to thrust from behind, or if it be occasioned by internal crystalline changes acting upon a rigid body; but it must be regarded as very remarkable, if the movement be that of a fluid body, no matter how viscous, for the length of the aclivity is sometimes several times the thickness of the ice. Crevassing, and other evidences of brittleness and rigidity, find a ready elucidation under the view that the ice is a really solid body at all times, and that its apparent fluency is due to the momentary fluidity of small portions of the mass assumed in succession as compression demands.

In addition to the considerations already adduced, it may be urged that a glacier does not flow as a stiff liquid, because its granules are not habitually drawn out into elongated forms, as are cavities in lavas and plastic lumps in viscous bodies. Flowage lines, comparable to those in lavas, are unknown in glaciers.

All this is strictly consistent with our primary thesis that a glacier is a crystalline rock of the purest and simplest type, and that it never has other than the crystalline state. This strictly crystalline character is incompatible with viscous fluidity.

BIBLIOGRAPHY

AGASSIZ, L. *Études sur les glaciers*, Neuchâtel, 1840.

CHAMBERLIN, T. C. "Presidential Address before the Geological Society of America." *Bull. Geol. Soc. Am.*, Vol. VI (February, 1895), pp. 199–220.

CHARPENTIER, J. DE. *Essai sur les glaciers et le terrain erratique du Basin du Rhône*, Lausanne, 1841.

CROLL, JAMES. "On the Physical Cause of the Motion of Glaciers." *Phil. Mag.*, Vol. XXXVIII (1869), pp. 201–6.

DEELEY, R. M. "A Theory of Glacial Motion." *Phil. Mag.*, Vol. XXV (1888), pp. 136–64.

DEELEY, R. M., AND FLETCHER, GEORGE. "The Structure of Glacier Ice and Its Bearings on Glacier Motion." *Geol. Mag.*, London, Decade 4, Vol. II (1895), pp. 152–62.

DRYGALSKI, ERICH VON. *Grönland-Expedition d. Gesells. f. Erdkunde zu Berlin 1891–93*, Vol. I (1897).

FORBES, J. D. *Occasional Papers on the Theory of Glaciers*, Edinburgh, 1859.

GRAD, CH. "La constitution et le mouvement des glaciers." *Revue sci.*, 1872.

HEIM, A. "On Glaciers." *Phil. Mag.*, Vol. XLI (1871), pp. 485–508; *Handbuch der Gletscherkunde*, 1885.

HOPKINS, W. "On the Theory of the Motion of Glaciers." *Phil. Trans.*, 1862, p. 677; *Phil. Mag.*, Vol. XXV (1863), p. 224.

HUGI, F. J. *Ueber das Wesen der Gletscher und Winterreise in dem Eismeer*, Stuttgart, 1842.

MALLET, R. "The Mechanism of Glaciers." *Jour. Geol. Soc.*, Dublin, Vol. I, p. 317.

"On the Plasticity of Glacier Ice." *Ibid.*, Vol. III (1845), p. 122.

"On the Brittleness and Non-Plasticity of Glacier Ice." *Phil. Mag.*, Vol. XXVI, p. 586.

MCCONNEL, J. C. "On the Plasticity of an Ice Crystal." *Proc. Roy. Soc.*, London, Vol. XLVIII (1890), pp. 256–60; Vol. XLIX (1891), pp. 323–43.

MOSELEY, H. "On the Cause of the Descent of Glaciers." *Brit. Ass. Rept.*, Part II (1860), p. 48; *Phil. Mag.*, Vol. XXXVII (1869), pp. 229, 363; Vol. XXXIX, p. 241; Vol. XLII, p. 138; Vol. XLIII, p. 38.

MÜGGE, O. "Ueber die Plasticität der Eiskrystalle." *Nachr. K. Ges. d. Wiss.*, Göttingen, 1895, pp. 1–4.

RENDU. "Théorie des glaciers de la Savoie." *Soc. Roy. Acad., Savoie, Mém. 1840* (in English, ed. by Geo. Forbes, London, 1874).

RINK, H. J. *Danish Greenland*, 1877.

THOMPSON, JAMES. "On the Plasticity of Ice as Manifested in Glaciers." *Roy. Soc. Proc.*, Vol. VIII (1857), pp. 455–58.

TYNDALL, J. *Forms of Water*, New York, 1872.

The Glaciers of the Alps, London, 1861.

TYNDALL, J., AND HUXLEY, T. H. "On the Structure and Motion of Glaciers." *Phil. Trans.*, Vol. CXLVII (1857), p. 327.

EXPLANATION OF PLATES

PLATE I

FIG. 1.— Snow crystals. (Photographed by W. A. Bentley.)

FIG. 2.—Terminal portion of Bryant Glacier, Inglefield Gulf, north Greenland, seen from the east, showing clean white ice above and stratified, discolored ice below, with a talus slope at the base. Height of vertical face above the gravel plain in front, about 140 feet. In this and

the following views the amount of débris in the discolored ice is much less than it appears to be, as it is greatly exaggerated to appearances by surface spreading.

FIG. 3.—Side view of middle portion of Bryant Glacier, showing the relation of the foliated débris-laden base of the ice to the clean white ice above.

FIG. 4.—Terminal portion of Bryant Glacier, seen from the west, showing the foliated structure of the basal ice.

PLATE II

FIG. 5.—A near view of the basal débris-bearing layers of the Bryant Glacier at a point where they are turned up into a nearly vertical attitude. The surface is much covered with débris derived from the ice above. This has been cut away from a portion to show the real amount of débris included in the ice, and its definite arrangement in layers.

FIGS. 6-8, text figures.

FIG. 9.—Portion of the end of South Point Glacier, Bowdoin Bay, Inglefield Gulf region, north Greenland, showing nearly pure ice above, but well foliated or stratified, and a bowlder-set layer below, from whose melting the talus slope of débris is formed. The ice of the glacier underlies this talus. The talus slope appears to be nearly or quite stationary, and the chief horizon of differential ice motion is probably at the bowlder-set layer at the top of the talus slope. The plain in front is formed of glacial wash.

FIG. 10.—Another portion of the end of South Point Glacier, showing the foliation of the ice brought out by shadows rather than by débris. The dark tract forming the lower part is ice covered with dirt from embraced débris. The base is a talus slope derived from this dark layer, but it is underlain by ice.

FIG. 11.—View of the edge of Krakokta Glacier—a tongue from the small ice-cap of Red Cliff Peninsula, Inglefield Gulf region, north Greenland, showing the strong individuality of the ice layers in the basal portion. The influence of débris in causing the layers to stand forth is here not very obvious, and the question of the individuality of motion of the layers is more open.

PLATE III

FIG. 12.—View of the terminal face of the Tucktoo Glacier—a tongue of the main ice-cap, Inglefield Gulf region, north Greenland, showing projection of the upper layers over the lower. The over-set of these layers is probably mainly due to the faster melting of the dirty ice below, and the fluting of the under surface of the projections is probably mainly due to the water from above, but the débris belt which is the primary cause, appears to have been formed by shearing. It is uncertain whether the shearing has immediately contributed anything to the projection.

FIG. 13.—Eastern edge of Bowdoin Glacier—a tongue of the main ice-gap, Inglefield Gulf region, north Greenland, showing contortion and shearing of the ice layers; movement from right to left. A ridge of rock juts out into the ice a little to the left of the part shown, and this is doubtless the occasion for the distortion and shearing.

FIG. 14.—Side view of the Gable Glacier—a terminal lobe of the main ice-cap of Greenland, Inglefield Gulf region, showing the taking on of basal load, and the development of shear-planes.

FIG. 15.—Striated surface of crystalline rock, near the end of Blasé Dale Glacier, Disco Island, Greenland.

FIG. 1

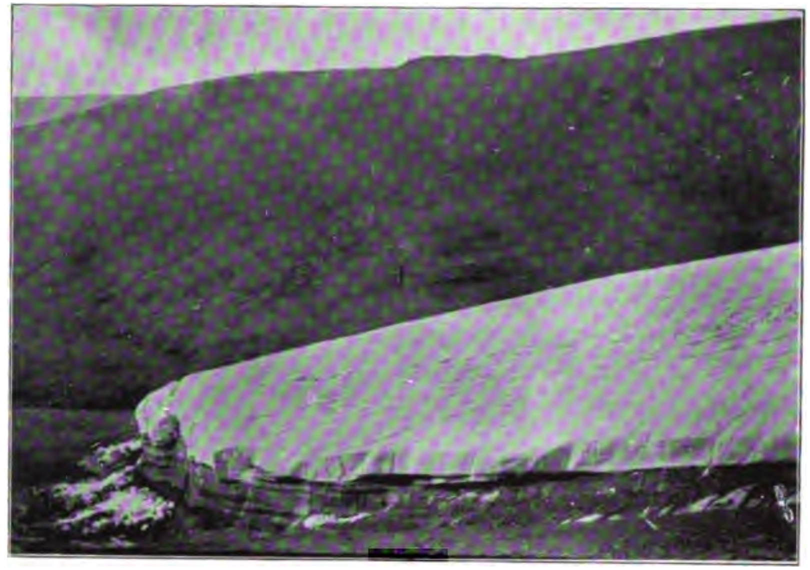


FIG. 2

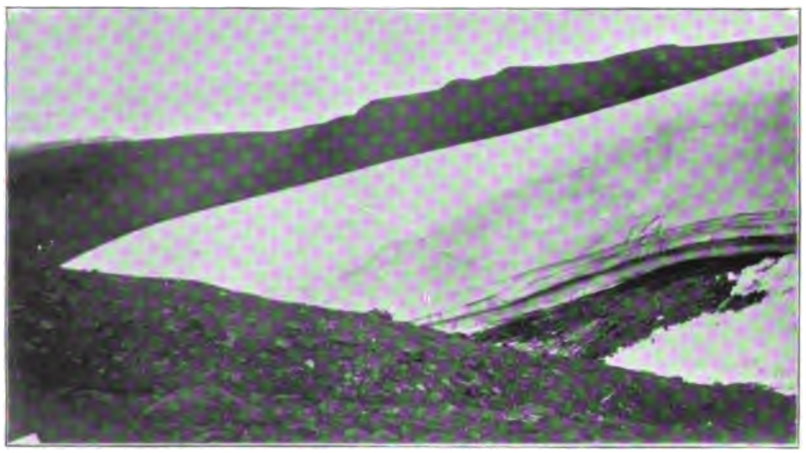


FIG. 3

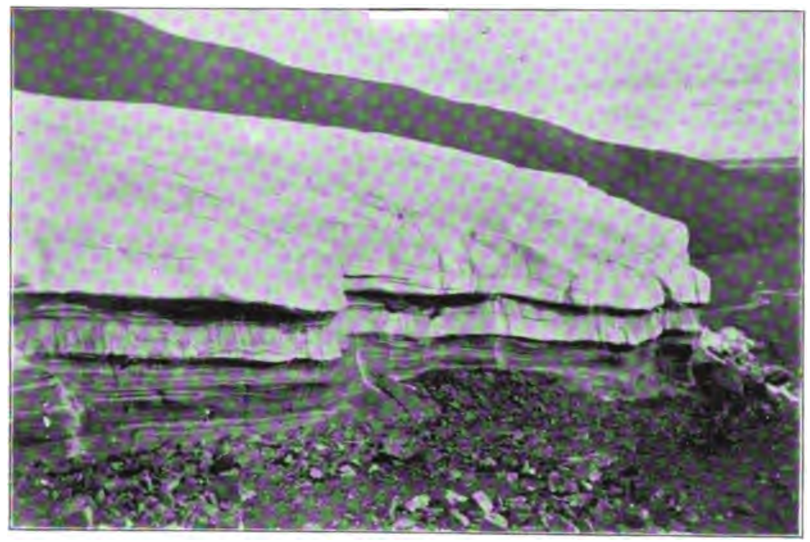


FIG. 4

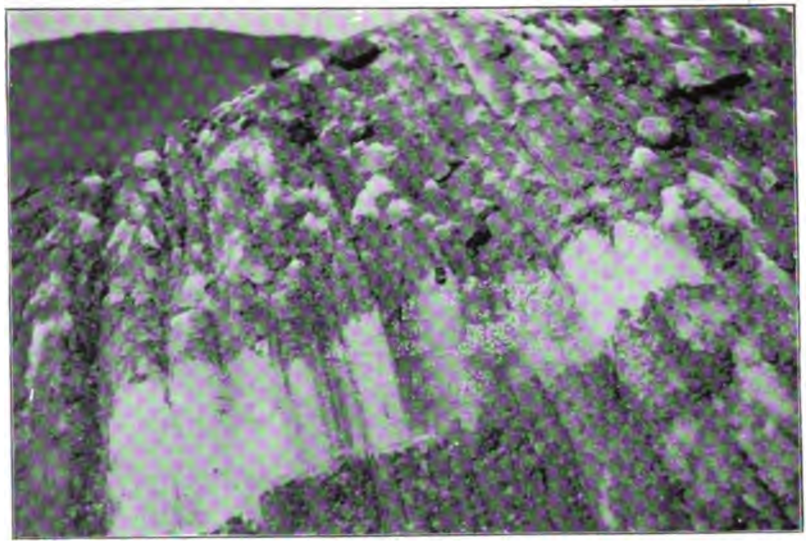


FIG. 5

FIG. 9

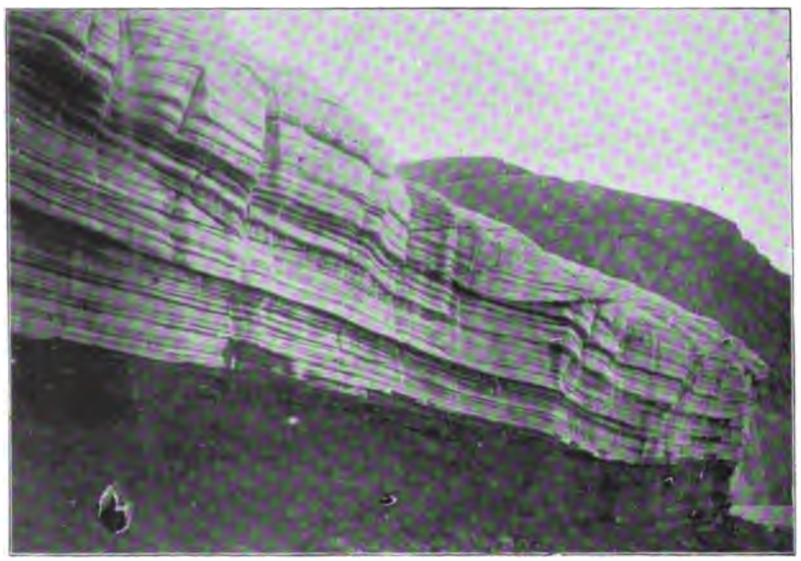


FIG. 10

FIG. 11

FIG. 12

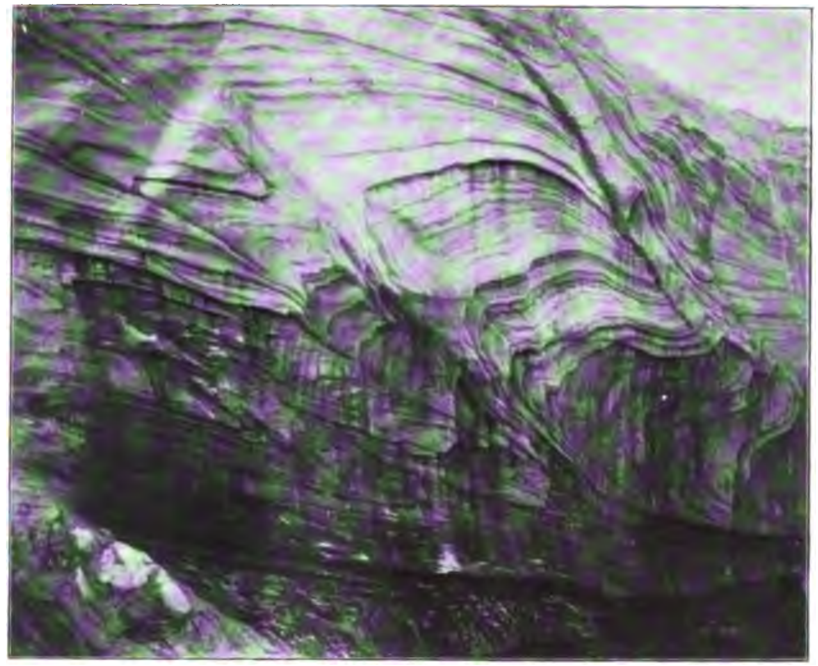


FIG. 13

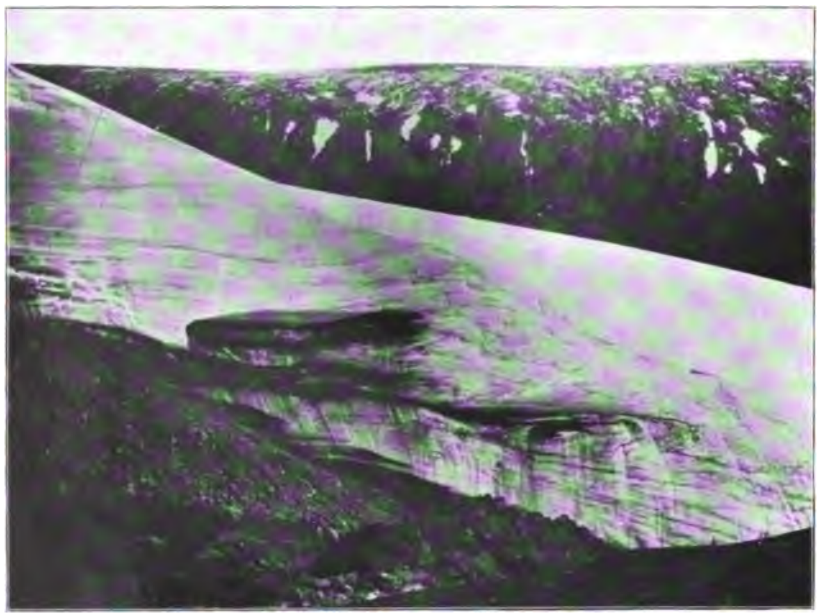


FIG. 14

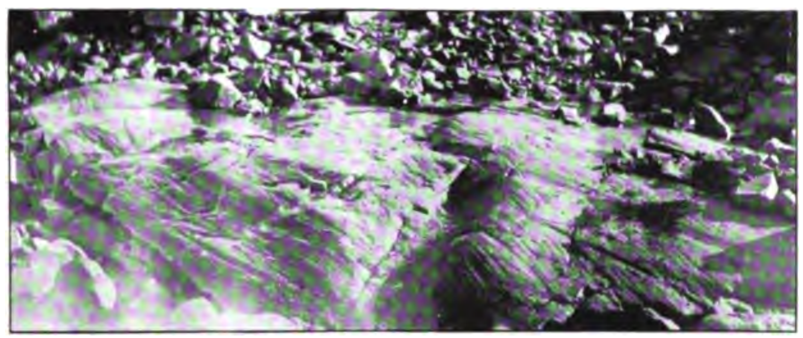


FIG. 15